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LOWER MANTLE P WAVE TRAVEL TIME UNDER ASIA

by

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Abstract

First arrival information reported by the International Seismological Centre and relevant to specific regions of the earth can be retrieved by computer from the magnetic tape versions of the ISC bulletins. A large number of P arrivals which reach their deepest point of penetration near the core mantle boundary was examined. The results suggest that the P velocity in the lower mantle under the Indian Ocean and Himalaya is low compared with the velocity under either Kamchatka or Mongolia. There is ambiguous evidence for a corresponding increase in attenuation of seismic energy.

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CHAPTER I : LOWER MANTLE HETEROGENEITIES

1. Introduction

The spherically symmetric 'average' earth has been well determined by the study of seismic waves generated by earthquakes and explosions, so recently more and more attention has been turned toward the study of regional deviations from such a spherically symmetric earth model. Lateral differences in the upper regions of the earth are evident from the studies of both surface and body waves and such differences may persist to depths of 700 km (Toksoz et al. 1967; Jordan, 1975a).

Evidence for the existence of lateral heterogeneities at greater depths is contaminated by the obscuring effect of structures in the upper mantle and crust. With increasing quantity and quality of data, as well as refinements in processing techniques, it may now be possible to assess the anomalies in the lower mantle. This thesis investigates this assumption for a data set relating to deep structures under Asia.

Considerable evidence has been presented by various authors in support of the existence of lateral variations in the lower mantle (Julian and Sengupta, 1973; Kanasewich et al, 1973; Jordan and Lynn, 1974; Dziewonski et al, 1977). It

has been suggested that these variations may be in some way related to mantle-wide convection, a possible driving mechanism of plate motion (Wilson, 1963; Morgan, 1972). Certainly, more knowledge of the lower mantle could provide clues to the nature of the earth's dynamic processes such as sea-floor spreading and continental drift.

The major problem facing all the studies of the lower mantle is the masking effect of structures in the crust and in the upper mantle. There are various corrections which may be applied to the raw data in attempt to eliminate the effect of the upper regions of the earth. These corrections will be discussed in Chapter 2.

2. P-wave Travel Times

Julian and Sengupta (1973) studied approximately 3300 p-wave travel times from 47 deep focus earthquakes. Deep focus events (depth between 450 km and 650 km) were chosen to avoid anomalous source effects (Sorrells et al, 1971). Constant station corrections, not dependent on the azimuth, were applied to reduce the contribution due to structures underneath the receivers. Epicentral parameters were recalculated to minimize gross errors due to mislocation. The resulting data were sufficiently accurate to resolve travel time anomalies in the order of 1.0 sec. Since the travel time of P waves passing through the lower mantle is of the order of 10 minutes, resolution in the order of 1 sec corresponds to an accuracy of 1 part in 600.

This study indicated a large number of anomalies in the depth range of 2600 km to the core-mantle boundary with velocity perturbations in the order of 1% or greater and physical dimension in the neighborhood of 1000 km. There was no apparent correlation between the lower mantle velocity anomalies and the global gravity anomalies or geoid height.

The travel time residuals was small; 0.16 sec or 0.23 sec delays were considered significant. Since the station corrections used in this study consisted of only a constant term and the large azimuthally dependent terms found by previous authors (Herrin and Taggart, 1968; Lilwall and Douglas, 1970) were not used, it was suggested that the small travel time residuals may be due to inadequate station corrections (Green, 1975).

Ellipticity of the earth corrections, which may be as high as 0.5 sec, are also important in this case and should be accounted for (Dziewonski and Gilbert, 1976).

3. Array Data

The travel time derivative with respect to epicentral distance Δ, is a useful parameter for the study of the mantle since at teleseismic distance ($\Delta > 30^{\circ}$) $dT/d\Delta$ data is independent of the origin time of the earthquake and is only slightly dependent on source effects such as radiation patterns, and source depth and location (Gutowski, 1974). A direct determination of this parameter can be obtained from

seismic array data.

Recently, a number of authors have presented evidence for compressional wave velocity anomalies in the lower mantle using information gathered from seismic arrays. Kanasewich et al (1973) analysed digital data from the University of Alberta seismic arrays: VASA and Peace, for events from the Tonga and Samoa islands. The phase velocity of the rays were found to have deviated up to 15% from results calculated using the Jeffreys-Bullen tables.

Since source effects are not significant, the contributions to the anomaly may come either from the region under the receivers or in the region near the midpoints of the rays, or both. Wright(1975) presented data from the Warramunga array in Australia and the Yellowknife array in Canada and pointed out that scattering below the array sites may be the cause of $dT/d\Delta$ anomalies observed by arrays. However, the use of large aperture arrays (180 km) will eliminate any coherent local anomalies. For the Kanasewich study, the travel times and phase velocities from events in other azimuths and distances are normal. This suggests that effects of structures under the array are negligible.

The most likely origin, in the sense that a small change has maximum effect, of the anomalous phase velocity is the lower portion of the ray path, near the core-mantle boundary. Various studies using PnKP phases (Buchbinder, 1972) and ScS phases (Hales and Roberts, 1970) have shown

that the radius of the core is unlikely to vary by more than a few kilometers. According to Kanasewich et al, the observed dT/dΔ for rays between Tonga and VASA would require a core depression of nearly 125 km; thus, bumps on the core-mantle interface can be ruled out as an explanation of the dT/dΔ anomalies.

Kanasewich et al concluded that the most likely cause of the observed dT/dΔ anomalies is a heterogeneous region of high velocity near the core-mantle interface in the region of bottoming points of the rays. On this interpretation the region under the Hawaiian Islands may relate to an upward plume of convecting material from the deep mantle.

Using data from the Large Aperture Seismic Array (LASA), Davies and Sheppard (1972) have also found anomalous values of dT/dΔ for rays passing under Hawaii. However, these authors do not place emphasis on the velocity inversion obtained from the dT/dΔ data. The velocities under Hawaii obtained from the LASA data are different from those obtained from VASA; however, the turning points of the rays indicate that the two arrays are sampling different regions of the lower mantle under Hawaii (Gutowski, 1974). A velocity distribution can be constructed to accommodate the results from both arrays.

4. Differential Travel Times

In the epicentral distance of 30° or greater, P and PcP

phases travel along approximately the same paths in the crust and in the upper mantle, and the paths differ only in the lower mantle. By taking the differential travel times such as PcP-P and ScS-S one can eliminate much of the anomalous effects of the crust and the uppermost mantle. For a spherically symmetric earth, the differential time should only depend on the epicentral distance. For the above reasons, differential travel times are especially suitable for the study of lateral heterogeneities in the lower mantle.

Jordan and Lynn (1974) carried out a study of ScS-S and PcP-P differential travel-times from two deep focus events in South America. The differential times were read from long-period seismograms with an accuracy of 1 sec or less. When they compared the results with the Jeffrey-Bullen tables, the authors found that the residuals showed a coherent azimuthal dependence which they interpreted as a region of high velocity in the lower mantle beneath the Caribbean. This is in agreement with the results of Julian and Sengupta.

The velocity anomaly was estimated to be greater than 1%. If the velocity contrast is due to a thermal anomaly, temperature differences of at least 200°c to 300°c would be required. The scale length of the lateral heterogeneities was estimated to be 500 km or greater. Jordan and Lynn suggested that this anomaly may be due to descending material in a convecting mantle.

Unfortunately differential travel time data must be obtained manually by careful analysis of many seismograms. The correct onset of PcP and ScS phases is often hard to determine. The S arrivals may sometimes be confused by precursors to S resulting from conversion of S to P at the base of the crust(Jordan and Lynn, 1974). In order to use these high resolution data, much time and manpower are required.

5. A Global Study

A study on a global scale using data from the ISC Bulletins was done by Dziewonski et al (1977). The detail of the ISC Bulletin will be discussed in Chapter 3. The precision of the ISC data is lower than those mentioned in the previous studies but the objective of Dziewonski's study is to determine whether the imprecision of this set of data can be compensated by the large quantity of observations.

Approximately 700,000 P-wave travel times were inverted to determine velocity anomalies at various depths of the mantle. The authors accepted a travel time observation only if

(1) The epicentral distance was greater than 27°.

(2) The event either was reported by at least 50 stations or had a magnitude greater than or equal to 4.8.

(3) The station has reported at least 100 events.

(4) The absolute value of the residual was less than 5

seconds.

The epicentral parameters were not recalculated. This assumes that any errors resulting from the use of J-B tables will be small for a good distribution of observations. The theoretical travel times were calculated using the PEM-C velocity model (Dziewonski, Hales and Lapwood, 1975). The travel times were corrected for the ellipticity of the earth using results published by Dziewonski and Gilbert (1976). In calculating the focal parameters, the ISC employed ellipticity corrections due to Bullen (1973). Dziewonski's ellipticity corrections agree with Bullen's to within 0.25 sec. The P wave travel times were also corrected for individual station anomalies. This azimuthally dependent correction is similiar to that used by Herrin and Taggart (1968a).

In determining the global distribution of p-wave velocity perturbations, the earth's mantle was modelled as four concentric homogeneous shells, of thickness from 670 to 800 km, divided by five latitudinal zones and six longitudinal zones. In the final study, the second shell was further subdivided into two layers. The velocity perturbation from a spherically symmetric velocity distribution was considered to be small and constant within each of the blocks. A least-square solution for the P-wave velocity perturbations, δv, in each of the 150 blocks was found using 693,495 observational equations.

Spherical harmonic coefficients for the distribution of the velocity anomalies were found using two methods. One method is by integration :

$$\begin{matrix} A_{ilm} \\ B_{ilm} \end{matrix} = \int_{\Omega} d\Omega \; \delta v_i(\theta,\phi) \; P_l^m(\cos\theta) \begin{matrix} \cos m\phi \\ \sin m\phi \end{matrix} \qquad (1\text{-}1)$$

Another method calculates the spherical harmonic coefficients as the least square solution to the system of equations:

$$\sum_{l=0}^{L} \sum_{m=0}^{l} A_{ilm} \overline{[P_l^m(\cos\theta)]_j} \; \overline{[\cos m\phi]_k} + \sum_{l=0}^{L} \sum_{m=1}^{l} B_{ilm} \overline{[P_l^m(\cos\theta)]_j} \; \overline{[\sin m\phi]_k} = \delta v_{ijk} \qquad (1\text{-}2)$$

where the bar denotes the average value of the functions in the particular zone. For each shell there are 5x6 values of δv_{ijk} and for L=3, there are only 16 coefficients to be determined. This overdetermined system acts as a high frequency filter and the δv_{ijk} are smoothed out by the low order continuous spherical harmonic functions. This form of representation is useful for comparison with gravity data.

Six well-defined velocity anomalies were found by the authors in the depth range of 2200 km to the core-mantle boundary: positive anomalies under Mongolia, the Central Pacific and the North Atlantic; negative anomalies under Central Africa, Australia and off the coast of Chile. The correlation between the velocity anomalies and the long wave-length gravity anomalies is negative; that is, regions

of high velocities are associated with low densities. There is no correlation between the distribution of velocity anomalies and the surface tectonic features and Velocity anomalies do not carry through vertically from layer to layer.

Compositional inhomogeneities and mantle wide convection might explain the negative correlation between the velocity and gravity anomalies. An upper limit on the dimension of the velocity anomalies in the lower mantle is placed at 5000 km.

The magnitude of the velocity anomalies is in the order of 0.07 km/sec or less which may be too small to be resolved by the P-wave data from the ISC Bulletins (See Chapter 3). The distribution of the anomalies in the lower mantle is not in agreement with the results found by Julian and Sengupta.

Dziewonski et al concluded that this set of data by itself is not adequate for global resolution of anomalies; however, it may resolve regional anomalies if a suitable distribution of sources and receivers can be found.

CHAPTER II : P WAVE TRAVEL TIME CORRECTIONS

In order to delineate any anomalies in the lower mantle through the use of seismological data, one must assess the effects of extraneous contributions such as source effects, structures under the seismic stations and deviation of the earth from a spherical configuration. Once the effects are known, it may be possible to eliminate them by applying appropriate corrections to the raw data.

1. Ellipticity of the Earth

The effect of earth's ellipticity of figure on travel times was first studied by Comrie, Gutenberg and Richter (1933). Bullen (1965) has presented a theoretical derivation of the effect.

The perturbation of the radial coordinate from a sphere due to ellipticity is given to sufficient accuracy by (Bullen 1965, 10-29):

$$\delta r = \epsilon r(1/3 - \cos^2\theta) \qquad (2\text{-}1)$$

where ϵ is the ellipticity at radius r, and θ denotes the geocentric colatitude of the point. The values of r range from -14 km at the poles to +7 km at the equater. Using equation (2-1), it is possible to obtain an expression for the correction to the travel time, accurate to first order, in the form (Bullen, 1965):

$$\delta T = \varepsilon_0 [(1/3 - \cos^2\theta)(\eta^2 - p^2)^{\frac{1}{2}}]_{0,\Delta} + \qquad (2\text{-}2)$$

$$1/p \int_0^\Delta \varepsilon (1/3 - \cos^2\theta) \eta^3 dv/dr \, d\theta$$

where ϵ_0 is the ellipticity of the earth's outer surface, p is the ray parameter and $\eta = r/v$. Using (2-2), Bullen constructed tables of ellipticity corrections for various phases. For practical applications, the approximate formula:

$$\delta T = f(\Delta)(H + h) \qquad (2\text{-}3)$$

is most frequently used, where H and h are the values of r at the epicentre and the observing station, and $f(\Delta)$ is a function of Δ alone given by tables (Bullen, 1939). For P waves, the ellipticity corrections range from -1.7 to +2.7 sec.

Dziewonski and Gilbert (1976) calculated ellipticity corrections based on an exact formulation of the ellipticity effect instead of the approximate formula given by equation (2-3). The difference between their results and Bullen's approximate formula are as large as 0.25 sec for P at 90° epicentral distance. For studies where a high degree of accuracy is required for the travel time, such accurate ellipticity corrections are essential. The ellipticity corrections are only slightly dependent on the velocity model as long as the model is 'close' to an acceptable standard earth model.

2. <u>Station Corrections</u>

Anomalous travel times are sometimes attributed to structures beneath the receiving stations. A number of workers have calculated station corrections to the travel times of P waves (Herrin and Taggart, 1968a; Cleary and Hales, 1966) by forcing travel time data to agree with a particular model of the earth. Often the resulting correction for a given station is small compared to the estimate of station variance.

Using data from 400 large earthquakes and 30 explosions, Herrin and Taggart (1968a) estimated azimuthally dependent station corrections for 321 seismological stations. A dependence on azimuth (Zij) of the form:

$$Tij = Ai + Bi * SIN(Zij + Ei) \tag{2-4}$$

is assumed for the source, and station i. Ai, Bi, and Ei define the ith station correction. These parameters were obtained through a least-square estimate using the travel time table of Herrin et al (1968) for 321 stations.

The results of this study showed that azimuthally dependent corrections up to 1 sec are required for some stations in the U.S. Corrections of similar magnitude are required for other stations in the world.

This traditional approach in calculating station corrections may not be legitimate in the analysis of P wave travel times in large epicentral distances. Empirically derived station corrections are biased by the distribution

of sources used to calculated them. They may contain as a result, precisely the effect one seeks, deep mantle velocity anomalies.

It is possible that in applying the station corrections to raw data, one may remove the effects of deep structure as well as effects of local station structures. Corrections such as those found by Herrin and Taggart are not available for many stations in the world.

3. Source Corrections

Seismically active regions are known to have strong lateral veloctiy variations. (Herrin and Taggart, 1968b; Sorrells et al, 1971). Beneath trenches the P wave velocity in the consumed lithosphere is several per cent higher than that of its surroundings. The presence of such a velocity contrast would shift the apparent position of the hypocentre as well as bias the travel times.

Once the structure near the source is known, the effect on travel times can be readily calculated by ray tracing. The main difficulty in applying the source correction is to decide what kind of structure exists near the source. Sorrells et al (1971) calculated the P wave travel times residuals caused by the presence of a thick slab of high-velocity material which dips steeply into the mantle. Comparisons were made with the observed data from LONGSHOT, an underground nuclear explosion detonated on Amchitka

Island on October 29, 1965.

Although excellent records of LONGSHOT were obtained at many stations, the errors in epicentral estimates were surprisingly large. Dubois (1966) attributed them to regional variations beneath the stations. Cleary (1967) and Herrin and Taggart (1968b) applied station corrections to the data and still found a strong azimuthal dependence in the data.

In order to explain the Longshot residuals Sorrells et al assumed the source to be near a bent slab 75 km thick, dipping in the northerly direction, reaching a depth of 300 km. The residuals predicted by this model are similiar to those observed. The residuals are small in the east-west direction, large and negative in the north direction and, positive going south from the source.

Another possible explanation for the source bias is anisotropy in the slab. Although no quantitative calculations are available for an anisotropic slab, experimental results suggest that anisotropy in the order of 10% is possible in preferentially oriented olivine (Morris et al, 1969).

Unfortunately most studies of the lower mantle do not consider source corrections quantitatively. Qualitative discussions do exist however.

CHAPTER III : THE DATA

1. The International Seismological Centre Bulletin

All the data used in this study were obtained from the magnetic tape versions of the International Seismological Centre (ISC) Bulletin. The contents of the ISC Bulletin have been stored on magnetic tape since 1964. In the time interval from 1964 to 1970, approximately 50,000 events were reported to ISC by approximately 3,000 Seismic stations located throughout the world. Of these stations only 1,400 are active and 500 observatories report most of the data.

The catalogue for the seven year period contains over 2 million P and PKP arrival times. This mass of seismic data is accessible to a computer for the study of the earth's lower mantle. These data are relatively inaccurate when compared with the more detailed data used by other studies, but the quantity of the data may compensate for their imprecision.

An explanation of the layout of the Bulletin may be found in the Introduction of the January, 1970 volume of the Bulletin. The earthquakes and explosions are listed in chronological order. Information on each event can be divided into 3 parts. The first part summarizes the preliminary estimates of the focal parameters as given by the various reporting agencies such as USCGS, MOS, and NOU.

The second part consists of revised estimates by the ISC. The International Seismological Centre recalculates focal parameters of the event by a revision program based on the Seismological Tables of Jeffreys and Bullen (1940). Only P, Pg, P* and Pn observations are used in the focal parameter calculations. Pg, P* and Pn are waves near the surface travel through superficial layers of the crust and are observed at distance less than 8° (Bullen, 1965).

Recalculated epicentral coordinates and origin times are given to one significant figure in the standard error. The focal depth is expressed both in fractions of an earth radius and in kilometers from the surface. Depths are shown to the nearest 0.0001R and nearest kilometer.

The third part consists of observational data reported by individual stations. The observational data consists of P and PKP arrival times. Some stations also give S and SKS arrival times but these are not included in the relocation calculations. Epicentral distances given are geocentric and are listed to the nearest 0.01 degrees. Station azimuths are given to the nearest degree. The P residuals are also listed to the nearest 0.1 sec

Reporting stations are identified by a three character station code and information on the stations such as geographic coordinates and elevations can be found in a station list compiled for each month's observations. On some of the earlier magnetic tapes (before July, 1965) this

station list is not available and a station list compiled by the United States National Earthquake Information Service is used. Station information can also be found in the Regional Catalogue of Earthquakes published monthly by the International Seismological Centre.

The problem of extracting data from the magnetic tape version of the Bulletin is non-trivial. There is a vast quantity of data and the formats lack consistency.

2. Resolution of the Data

Aside from anomalous structures in the earth, there are two major sources of error in any seismic data : (a) errors in estimating the arrival times from the records and (b) errors in estimating the source parameters. Freedman (1968) gave a detailed discussion of errors associated with seismological data. The following is a summary of these errors and the procedures used in this study to minimize their effects on the resolution of the data obtained from the ISC Bulletin.

(a) Arrival Time Errors

(i) miscount - Miscounts of time marks on the seismogram results in errors in multiples of seconds, minutes or even hours. Truncation of data is one method of eliminating some of the miscount errors. In this study, residuals of absolute value over five seconds are considered as gross errors and are hence discarded.

(ii) misidentification - The first arrival may be missed due to a high level of background noise. Systematic errors may result for a particular station due to the first arrival selection 'habits' of its operator (Herrin and Taggart, 1968a). Gross misidentification of phases may be eliminated by the truncation method mentioned above. To further compensate for possible errors from misidentification an algorithm that weights large late arrivals was used to select first arrivals when calculating the theoretical travel times (Roebroek and Nyland, 1975).

(iii) instrumental errors - Variation in instrumental response, paper speed fluctuations, etc. are examples of instrumental errors. These errors are much smaller than the others and may be ignored.

After truncation of data and elimination of gross misidentifications, a conservative estimate of the errors in the estimate of arrival times can be placed at 0.5 sec for P waves (Freedman, 1968).

(b) Errors in Source Parameters

Errors in the estimates of origin time and hypocenter of an earthquake are more difficult to pinpoint. Since the estimates of the source parameters are complicated functions of the arrival times, their errors are dependent on the errors of

the travel times. In calculating the epicentral parameters, a least-squares fit of the observed travel times to those calculated from a spherically symmetric earth model is made. The choice of this model must contribute errors to the source parameter estimates. The ISC uses the Jeffreys-Bullen tables for its epicentral calculations. The Jeffreys-Bullen tables are sufficiently accurate for the average earth that, given a large number of observations, the errors should not be significant (Dziewonski, 1977).

For each event, the ISC also lists the estimates of the errors for the various focal parameters such as origin time and depth. By studying these errors one may be able to determine the resolution of the ISC data. Also listed is the deviation (from the J-B tables) for a single observation.

The average of the standard deviations of a single observation was calculated for 76 events in the Philippine region. All events are from the month of January, 1970.

The average standard deviation was 1.62 seconds. Even though this standard deviation is with respect to the J-B tables, it still gives an estimate on the intrinsic resolution of the data. With a large number of observations, the resolution of the data may improve; however, it is difficult to determine

quantitatively how the accuracy changes with the number of observations due to uneven distribution of events and receivers. It is safe to say that, after truncation, residuals with magnitude greater than 2 seconds are significant. Residuals less than 1 second may result from causes other than anomalies in the earth.

3. The Source Regions

Table 3-1 summaries the seismic regions from which all the events for this study were chosen. The source regions may be divided into two main groups : the western margin of the Philippine Sea Plate and the Sunda Archipelago.

(a) The Phillipine Sources

For convenience, the first group of sources shall be called the Philippine sources. Much of the western margin of the Philippine Sea plate appears to be in a transition stage of a developing plate boundary that consists of deformation zones for the accommodation of plate motion (Rowlett and Kelleher, 1976). Four segments of the boundary can be identified : the Ryukyu arc, Taiwan, the Luzon strait and the Philippine Islands (see figure 3-1).

Katsumata and Sykes (1969) observed typical features of a subducting margin along the Ryukyu arc with earthquakes concentrated in a planar zone

dipping 35° - 45° northwest to a depth of about 280 km. Rowlett and Kelleher (1976) found that along the portion of the Ryukyu arc between Okinawa and the Japanese island of Kyushu, occurrence of large earthquakes is very infrequent.

Southwest of the Ryukyus, in the vicinity of Taiwan, a collisional boundary between the Asian continental lithosphere and the Luzon arc has been proposed (Chai, 1972). This is a fairly active region in terms of seismicity.

Earthquakes observed around the Luzon Strait are generally shallow and of moderate magnitutes ($M < 6.9$). This diffuse zone of seismic activity suggests tectonic strain releases regionally rather than along a well-defined boundary (Rowlett and Kelleher, 1976).

Earthquakes in the Philippine Islands are also shallow and are distributed in a diffuse pattern throughout the Islands. Many large earthquakes are clustered along or near the Philippine Trench and the Philippine Fault.

Also grouped with the Philippine sources are earthquakes from the Borneo region just south of the Philippine Islands.

(b) The Sunda Sources

The second group of earthquakes, called the

Sunda sources, is much better defined than the Philippine sources. It consists of events from the Western Sunda Arc and the Eastern Sunda Arc with the Sunda Strait acting as a physiographic break as well as a tectonic break (Fitch, 1972).

The Western Sunda Arc is a region of shallow underthrusting of the oceanic lithosphere associated with the Indian Ocean-Australian plate. Seismicity generally terminates at the depth of 200 km and the angle of dip of the descending slab is not well known.

There is further evidence of the tectonic break from the concentration of shallow earthquakes in the Sunda Strait region. East of the Sunda Strait is the Eastern Sunda Arc where earthquakes as deep as 600 km are reported. A dominant bathymetric feature of this region is the Java Trench. Seismicity profiles indicate a dip of 30° (Fitch, 1970).

The Sunda Archipelago represents a tectonic feature that is approximately perpendicular to most of the features along the western margin of the Philippine Sea plate; thus, the two groups of sources may aid in the study of source effects.

4. <u>Time Period of the Events</u>

Unlike Dziewonski et al (1977) in their global study,

where events from all the seven available years were used, the events in this study were selected only from the years of 1969 and 1970. It was found that additional data did not change the general results obtained from the data of two years. It was concluded that the short time period was not a fundamental limitation in the data and that the increase in computing expenditure was not justified by a small increase in information if a longer time interval was searched for data. For the years 1969 and 1970, approximately 14,000 P wave arrival times were obtained in the epicentral distance range of 60° to 110° for events in the seismic regions listed in table 3-1.

All observation data were taken as they are reported in the ISC Bulletin. No recalculation of the epicentral parameters was made.

Figure 3-1 : The major tectonic features of the Philippine source region: Longitudinal Fault of Taiwan (L.F.), Philippine Fault (solid line), active volcanoes (triangles) and trenches (saw teeth on upper plate). The 100-fm (183m) contour represents the approximate boundary of the continental margin. (After Rowlett and Kelleher, 1976)

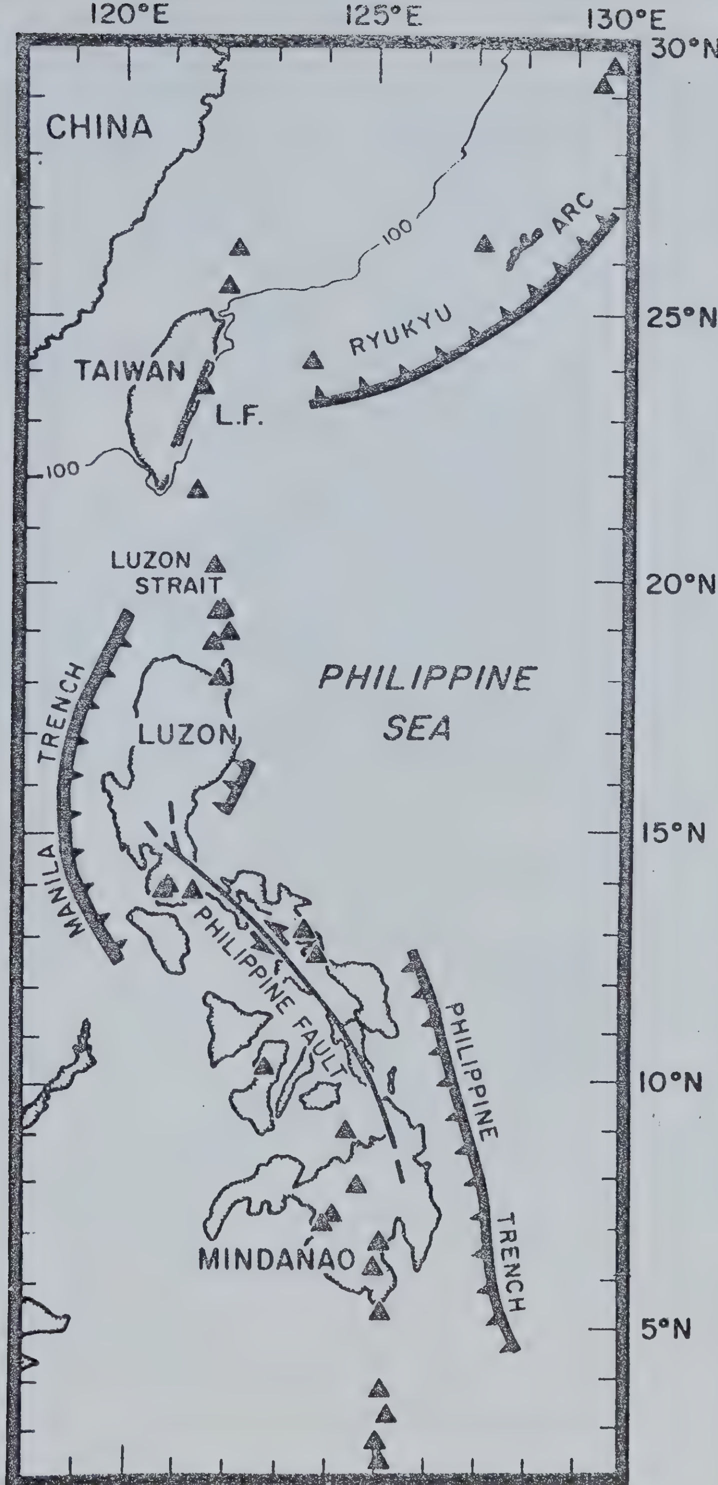
120°E
125°E
130°E
30°N
25°N
20°N
15°N
10°N
5°N
CHINA
100
RYUKYU
ARC
TAIWAN
L.F.
100
LUZON
STRAIT
PHILIPPINE
SEA
LUZON
TRENCH
MANILA
PHILIPPINE FAULT
PHILIPPINE
TRENCH
MINDANAO

Figure 3-2 : The major tectonic features of the Sunda Arc source region. Shading and striping mark regions of known or suspected extension and shortening, respectively. Directions of relative motion between the major plates are given by solid lines (Morgan, 1972) and dashed lines (Le Pichon, 1968). (After Fitch, 1972)

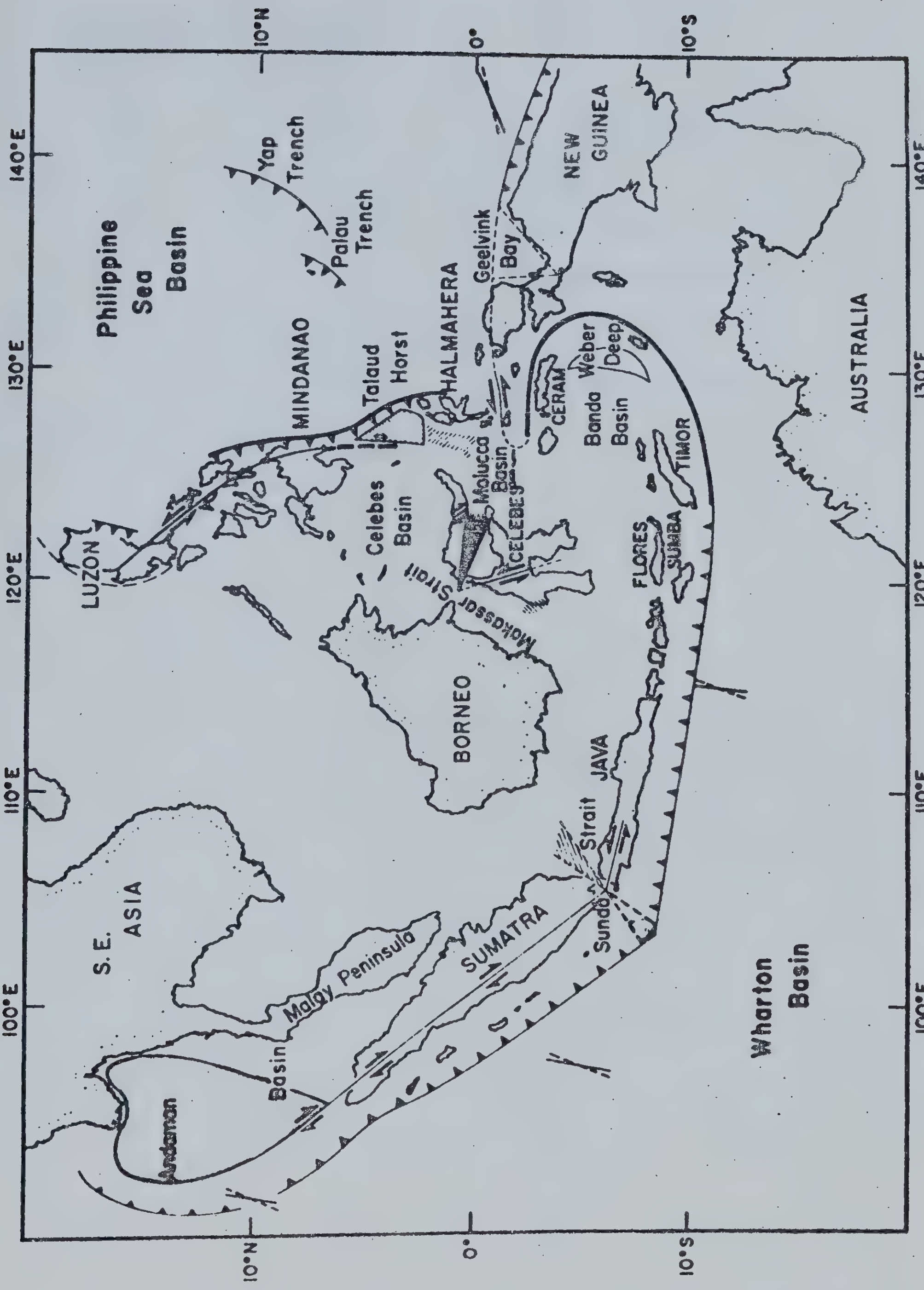
Philippine Sea Basin
Yap Trench
Palau Trench
MINDANAO
Talaud Horst
HALMAHERA
Geelvink Bay
NEW GUINEA
Weber Deep
CERAM
Banda Basin
TIMOR
AUSTRALIA
LUZON
Celebes Basin
Molucca Basin
CELEBES
Makassar Strait
FLORES
SUMBA
BORNEO
JAVA
Sunda Strait
SUMATRA
S.E. ASIA
Malay Peninsula
Andaman Basin
Wharton Basin
10°N
0°
10°S
100°E
110°E
120°E
130°E
140°E

TABLE 3-1

SEISMIC REGION NUMBER	SUB-REGION INDEX	NAME OF REGION
20	238-241	RYKUKYO ISLANDS
21	242-247	TAIWAN
22	248-260	PHILIPPINES
23	261-272	BORNEO-SOLAWESI
24	273-293	SUNDA ARCHIPELAGO

CHAPTER IV : P WAVE TRAVEL TIME UNDER ASIA

1. Introduction

One of the most prominent velocity anomalies found by Dziewonski et al (1977) in the depth range of 2200 km to the core-mantle boundary is a high velocity region under Mongolia. A velocity perturbation in the order of 45 m/sec was found in this region. Using the same data base, a study on a much more restrictive and regionalized scale was made on the Mongolian region to see if such a result is repeatable. Neighbouring regions in the lower mantle were also examined for the purpose of comparison.

2. Data Selection Procedures

To sift through the huge quantity of data available on the magnetic tape version of the ISC Bulletin, an improved version of a computer program package, originally written by E. Nyland, E. Roebroek and L. Rossell at the University of Alberta, was used in an automated data search.

The first part of the program, called PISC1, is a read and write operation that picks out relevant information from the tapes according to the restrictions defined by the input parameters to the program. The following is a description of the necessary input parameters to PISC1.

(a) Time interval to be searched: Time intervals to be searched for data are defined by months and years.

Each 2400 foot magnetic tape contains earthquake information for a six month period and the tape is positioned to the correct month before the data reading phase of the program begins.

(b) Region numbers : Seismic regions of the world are assigned unique region numbers (Flinn and Engdahl, 1965) according to their geographic locations. This number is listed for each event and it is a useful index for earthquake searches. A list of these numbers corresponding to their geographical area may be found in the Regional Catalogue of Earthquakes.

(c) Source boundary coordinates : This input further defines the limits of the location of the event by latitude and longitude coordinates.

(d) Limits of epicentral distance : Arrival times of P phases are extracted only from stations with epicentral distance falling within this limit. This indirectly defines the depth of penetration of the P phases.

The output of PISC1 consists of the following : seismic region number of the earthquake, latitude and longitude of the epicentre, depth of the epicentre, origin time of the earthquake, station code of the station, arrival time at the station, epicentral distance, and the station azimuth. This output will be further processed by the second part of the program package, named PISC2, which calculates travel time residuals for specific regions of the earth.

3. Travel Time Residual Calculations

The theoretical travel times for each observed phase were calculated by a subroutine which was written by M.J. Randall, M. Shimshoni, J. Gardner and E. Nyland at UCLA. The velocity model used for the calculated travel time was the Jordan-Anderson Model B1 for P waves (figure 4-1, Jordan and Anderson, 1974).

Model B1 was inverted from an extensive set of free oscillation data and differential travel-time data. The free oscillation data give good parameterization of the earth's crust and upper mantle while the differential travel-times provide high resolving power for the lower mantle. Absolute travel times and $dT/d\Delta$ data were not used in the inversion.

This model is characterized by the lack of a low velocity zone for compressional waves and two discontinuities in the upper mantle transition zone at depths of 420 and 671 km. The radius of the core is fixed by PcP-P times at 3485 km, 12 km greater than the value obtained by Jeffreys and Bullen.

The International Seismological Centre calculates epicentral parameters by using the Jeffrey-Bullen Model but a different model, the B1 model, is used in calculating the differences between observed and theoretical travel time for the following reasons:

(1) New observations, especially ultra-long period seismological data, indicate a need for revision of

the J-B tables (Herrin et al, 1968; Hales and Roberts, 1970). The B1 Model is expected to fit observations from an average earth to a better accuracy than the J-B Tables.

(2) The errors resulting from using two different velocity models in the two stages of the data processing are expected to be small compared to other errors; thus, no recalculation of the epicentral parameters is required.

(3) Model B1 is only used as a reference model. Emphasis will be placed on the relative behaviour of the residuals from different regions of interest rather than on the absolute travel time residuals, ie, to see if certain region is faster or slower than other regions.

All calculated travel times were corrected for ellipiticity of the earth by using the approximate formula (See chapter II):

$$\delta T = f(\Delta)(H + h) \qquad (2\text{-}3)$$

The exact corrections of Dziewonski and Gilbert (1976) were not used because the difference between the two corrections is not significant for this set of data.

No source corrections were applied to the data. Since for many stations used in this study there are no well-determined station corrctions available, no station

corrections were applied. Instead, observations from a wide distribution of stations were used to reduce any coherent station effects.

4. Results of a Regional Study

Teleseismic rays spend 25% of their travel time in the lower 10% of their paths (Julian and Sengupta, 1973); therefore, the contribution to the total travel time from the neighbourhood of the rays' bottoming points is significant. If one was able to eliminate upper mantle and crustal anomalies, the effects of the lower mantle could be studied by partitioning the observed phases into groups according to their bottoming points. This is the approach used by Julian and Sengupta in their travel times study where they arbitrarily assigned the central 30° of the ray path as the contributing region to any anomalous residuals.

In this study, the four regions of interest (Table 4-1) were defined by four-sided polygons and the midpoint of each observed phase was calculated. Then the observed phases were either accepteded or rejected by whether their midpoints were located within a region of interest.

(a) Philippine Sources

(i) Kamchatka Region

Figure 4-2 shows the earthquakes and stations used for the study Kamchatka. Also shown are the midpoints of the observed phases relevant

to this region. The travel time residuals of these phases with respect to the Jordan-Anderson Model are plotted on figure 4-3. The epicentral distance range of the data is 60° to 105°, corresponding to a depth range of approximately 1500 km to the core-mantle boundary for the deepest point of penetration of the rays.

A least-square fit polynomial is plotted through the data points. The maximum degree of the polynomial is three. The objective in using such a curve is not really to imply that the residuals are such a mathematical function of the epicentral distance; rather, it is intended to aid in visualizing the qualitative nature of the dependence of the residuals on delta.

How well the curve describes the behaviour of the data points can be indicated by a quantity called the coefficient of determination. The total variation of the residuals is defined by:

$$\sum_i (t_i - T) = \sum_i (t_i - f_i) + \sum_i (f_i - T) \qquad (4-1)$$

where t is the travel time residual, T is the average of the residuals and f is the estimate of the residual based on the assumed function. The first term on the right of (4-1) is called the unexplained variation while the second term is

call the explained variation, so called because the deviations (f_i-T) have a definite pattern while the deviations (t_i-T) behave in a random or unpredictable manner.

The ratio of the explained variation to the total variation is called the coefficient of determination, CD. If the total variation is all unexplained, the ratio is zero and if the total variation is explained, the ratio is one. The IMSL Library subroutine that was used in calculating the least-square fit curve controls the degree of the fit by using the coefficient of determination as a parameter (Forsythe, 1957).

Table 4-2 lists the coefficients of determination of the polynomial fit up to the fourth order. The value of CD computed for each case measures the relationship between the residuals and delta relative to the type of equation assumed. For example, if CD is nearly zero for the linear fit, it means that there is no linear correlation between the residuals and delta. The value of CD thus determines the significance that one can attach to the curve.

The generally small values of the coefficients of determination indicate that it is not reasonable to place too much faith in the

cubic curve; however, visual inspection shows that the cubic curve fulfils its intent as a rough representation of the data. No further use is made of the curve.

The length of the error bars here, as elsewhere, is two times the standard error of the data set with respect to the polynomial. The interval of one standard error above and below the polynomial represents the 68.3% confidence interval, that is, 68.3% of the data points are expected to fall within this interval. The size of the error bars gives useful information on the scatter of the data. In case of the Kamchatka data, the standard error is 1.1 sec and the total number of observations is 598.

In their study of deep focus earthquakes, Julian and Sengupta found the velocity of the lower mantle beneath the Kamchatka region to be low with respect to the J-B tables. Since the calculated travel time from the Jordan-Anderson model is approximately 2.5 sec earlier than that from the J-B tables at the epicentral distance of 60° and the difference between the two models decreases to approximately 1.0 sec at 90°, slow arrivals with respect to the J-B tables would be much slower when compared with the J-A model.

No signigicant anomaly is evident in the travel time residuals shown in figure 4-2 except for a positive 1.5 sec. in the 60° to 70° range.

(ii) Mongolian Region

Figure 4-4 shows the phases and stations used for the plot of travel time residuals in figure 4-5. As with the Kamchatka region, the events used were from the Philippine sources. Notice in figure 4-4 that some of the events along the Ryukyu arc were not detected by this set of stations whereas the North American stations did observe these events. This may confirm what Rowlett and Kelleher have found, that the earthquakes occurring in this neighbourhood are of small magnitude and they are detected by the stations located in Alaska more readily than by the European stations.

The graph on figure 4-5 shows a large concentration of data near the 90° range and the overall zero residual is evident. The positive trend beyond 100° is probably due to the exaggeration of the cubic polynomial representation; nevertheless, the majority of the residuals are positive in the epicentral distance of 100° and greater.

Dziewonski et al have found this region of the lower mantle from 2200 km depth to the core-mantle boundary to be fast. This depth range corresponds to the delta range from 75° to 105°, but figure 4-4 shows that this set of data indicate normal behaviour with respect to the Jordan-Anderson model. There are 769 points on this graph with standard deviation of 1.4 sec.

(iii) Himalaya Region

The third region sampled by the Philippine sources is the Himalayas (see figure 4-6). This includes the collision boundary between the Indian plate and the Asian plate. It would be of interest to examine the lower mantle under the Himalayas to see if there is any anomaly that would correlate with this surface feature.

Figure 4-7 shows the 212 travel time residuals obtained for phases that bottomed under Himalaya. The standard deviation increased to 1.5 sec indicating a greater degree of scattering, evident in the 90° to 100° delta range. The polynomial shows a definite positive bias in the residuals ranging from a maximum of 2.5 sec near 70° and decreasing with epicentral distance to 1.0 sec near 100°.

(iv) Indian Ocean Region

The geoid has a minimum in the Indian Ocean just south of India. This region is sampled by phases, received by stations in Africa, from the Philippine source earthquakes. The available observations are greatly reduced in number even though the same set of earthquakes used in the previous three regions is searched for P arrival times. Most of the events detected by the eight African stations are located in the Philippine Islands area (see figure 4-8).

Figure 4-9 shows the 46 travel time residuals plotted against the epicentral distances. The standard deviation is 1.6 sec Due to the distribution of stations and events, there were no observation in the delta range of 60° to 80°. Even though the number of observations is much smaller than at other regions, the positive bias of the data is quite pronounced. Observed travel times were delayed up to 3 sec compared to the calculated travel times.

(b) Sunda Sources

A second group of earthquakes, the Sunda Sources, was used to sample the Mongolia, Himalaya, and the Indian Ocean regions. The midpoints of the phases from the Sunda Sources

do not fall in the Kamchatka region, hence, no comparison can be made for this region from the use of two different groups of sources.

(i) Mongolia Region

The travel time residuals for the Sunda Source are plotted in figure 4-11. The polynomial indicates that the average of this set of residuals is again near zero for all available delta ranges. Most of the observations were clustered around the epicentral distance of 100°.

(ii) Himalaya Region

The polynomial in figure 4-13 is much closer to the zero line than that of figure 4-6. The difference is especially noticeable around the 70° delta range, where the residuals have gone from a positive 2.5 sec for the Philippine source to zero for the Sunda source. The scatter of the data is also less.

(iii) Indian Region

Figure 4-13 shows that only two African stations were reporting the P phases from the Sunda sources. The small number of observations again indicate a definite delay in the travel

time of P phases in the order of 4 sec.

5. Interpretation of the Results

Even though the Jordan Anderson B1 model satisfies a large number of seismological observations, there may be systematic errors inherent in the residuals of all regions of the earth due to the usage of this, or any reference model. However, the magnitude of this error should be small (Jordan and Anderson, 1974) and for the purpose of the interpretation, a particular set of residuals may be choosen as a reference, thus minimizing the ambiguity of an inadequate reference velocity model.

Of the seven sets of ray paths, the Philippine-Kamchatka ray path was choosen as a reference. The residuals for this path are close to zero for most epicentral distances. There is a fairly even distribution of available observations throughout the delta range of interest (see figure 4-3).

Using the residuals of the Philippine-Kamchatka ray path as a reference, the behaviour of the other six sets of residuals can be disccussed. A positive bias in the residuals for the delta range of 60° to 80° is characteristic for most of the ray paths where the configuration of earthquakes and stations was able to provide observations at this delta range. This anomaly may be explained by accepting the hypothesis that the velocity

model for this depth range (1500 km to 2400 km) is probably too fast, resulting in a positive travel time residual in the order of 1 sec.

(1) Philippine-Mongolia Ray Path

For most of the epicentral distances, especially between 80° and 100°, the residual curve for the Philippine-Mongolia path (figure 4-7) is very much the same as that of the Philippine Kamchatka path. The strong upward tilt of the curve beyond 100° is probably exaggerated by the cubic representation; nonetheless, the predominately positive residuals may indicate a low velocity region in the depth range of 2800 km to the core mantle boundary.

Dziewonski et al have found that the lower mantle velocity under Mongolia is 0.03 km/sec higher than that under Kamchatka. They have suggested a maximum linear scale for the anomaly of 5000 km or less. This would produce a difference in the travel time residuals between the two regions of about 0.8 sec in the delta range of 75° to 105°. Such a difference is not evident from the two sets of residuals in figure 4-5 and figure 4-3. The standard error of 1.4 sec for mongolia and 1.1 sec for Kamchatka, suggests that a difference of 0.8 sec may be undetectable.

(2) Sunda-Mongolia Ray Path

In contrast with the Philippine-Mongolia path, the number of available observations for the Sunda-Mongolia path is small (see figure 4-4) and the residual curve is a constant -0.9 sec throughout all delta range. This set of residuals does appear to indicate that the Mongolia region has a higher velocity than the Kamchatka region by the appropriate amount.

This negative residual may also be the effect of a down dipping slab of high velocity material near the source. Sorrells et al have shown that it is possible to have residuals due to structures under the source up to -1.9 sec in the azimuth such as that between the European stations and the Sunda Arc. Source effect is certainly a plausible explanation for the negative residuals shown in figure 4-11 in view of the well developed island arc system of the Sunda sources and the evidence of the existence of a down dipping slab here. Such an azimuthally dependent source effect may not be present in the Philippine sources since the locations of the epicentres do not suggest a systematic arc structure (see figure 4-10).

(3) Philippine-Himalaya Ray Path

The travel time residuals (figure 4-7) for this

ray path are mostly positive. The observed travel times in the delta range between 80° and 100° are up to 2 sec higher than those for the Philippine-Kamchatka ray path and the Mongolia ray paths. The wide distribution of receiving stations is not expected to produce any coherent station effects in the travel times. Again, source effects from the Philippine sources are not likely to be important due to the fact that major seismic activity here has not yet consolidated along any well-developed transform and subduction boundaries (Rowlett and Kelleher. 1976);rather, earthquakes seem to be clustered in an uneven distribution.

This set of residuals suggests that the lower mantle under the Himalaya may have a lower velocity than the lower mantle under either Kamchatka or Mongolia.

(4) Sunda-Himalaya Ray Path

The travel time residuals for the Sunda-Himalaya ray path (figure 4-13) are strikingly different from those of the Philippine-Himalaya ray path. The residual curve is up to 2.5 sec lower than the Philippine-Himalaya residuals. As with the difference between the Sunda-Mongolia and Philippine-Mongolia residuals, the difference between the two sets of Himalaya residuals may be attributed to the source

effect of the Sunda Arc subduction zone. If this is true, then the Sunda-Himalaya residuals also indicate a lower velocity under the Himalaya.

(5) Philippine-Indian Ocean Ray Path and Sunda-Indian Ocean Ray Path

There are two significant differences between the residuals for the Indian Ocean region and the residuals of the other three regions: one, the number of observations is much less than can be accounted for by the reduction in the number of seismic stations in Africa; and two, of the few observations available, all indicate a significant delay in the travel time of P waves.

According to Sorrells' model, the Sunda source effect would be negligible for the Sunda-Indian Ocean ray path. In the direction parallel to the strike of the subduction zone, residuals due to source structure are nearly zero. The fact that only two seismic stations provided all the observations for the Sunda-Indian Ocean path means that the structure under the stations may be important. The argument that a wide distribution of a large number of seismic stations would eliminate any coherent receiver effects is no longer applicable.

The number of stations for the Philippine-Indian

Ocean path increases to eight and again the residuals show large delay in the observed travel times. If structures under the stations are responsible for the anomalies, then such structure must extend throughout a large area of the African continent since the eight African stations are located over a wide region (see figure 4-8).

It is unlikely that any crustal structure can account for the large delay of about 4 sec. Over a linear dimension of 200 km say, such delay would require a velocity contrast in the order of 10% which is not likely over such a wide region of the African continent.

A possible explanation for the large delay in the travel time of P waves and the reduced number of observations is the existence of an absorptive low velocity region in the vicinity of the lower mantle under the Indian Ocean where the rays reach their deepest point of penetration. This low velocity region would coincide with the large geoid low just south of India which has been suggested by Gough (1977) as the location of a global convection upcurrent. Gough's suggestion is based on the asymmetric shape of the geoid and the high northward relative velocity of the Indian plate. Such an anomalous region would tend both to attenuate and slow down the propagation of seismic energy.

This correlation between a low velocity region and a negative gravity anomaly in the lower mantle is not consistent with the results of Dziewonski et al (1977). In the Julian-Sengupta study, there are no observations available for this region of the lower mantle. However, the small number of observations presented here makes such a comparison inconclusive.

6. Conclusion

In this study, the data from the International Seismological Centre Bulletin were used to examine four regions of the lower mantle. The only evidence of lower mantle inhomogeneities was found in the Indian Ocean region and the Himalaya region. The interpretation of the results is not unique and comparisons with other studies are not conclusive. However, the International Seismological Centre Bulletin appears to be a good source of information for the initial study of a particular region. With the aid of the computer program, such a preliminary study can be done with a reasonable expenditure of time and manpower.

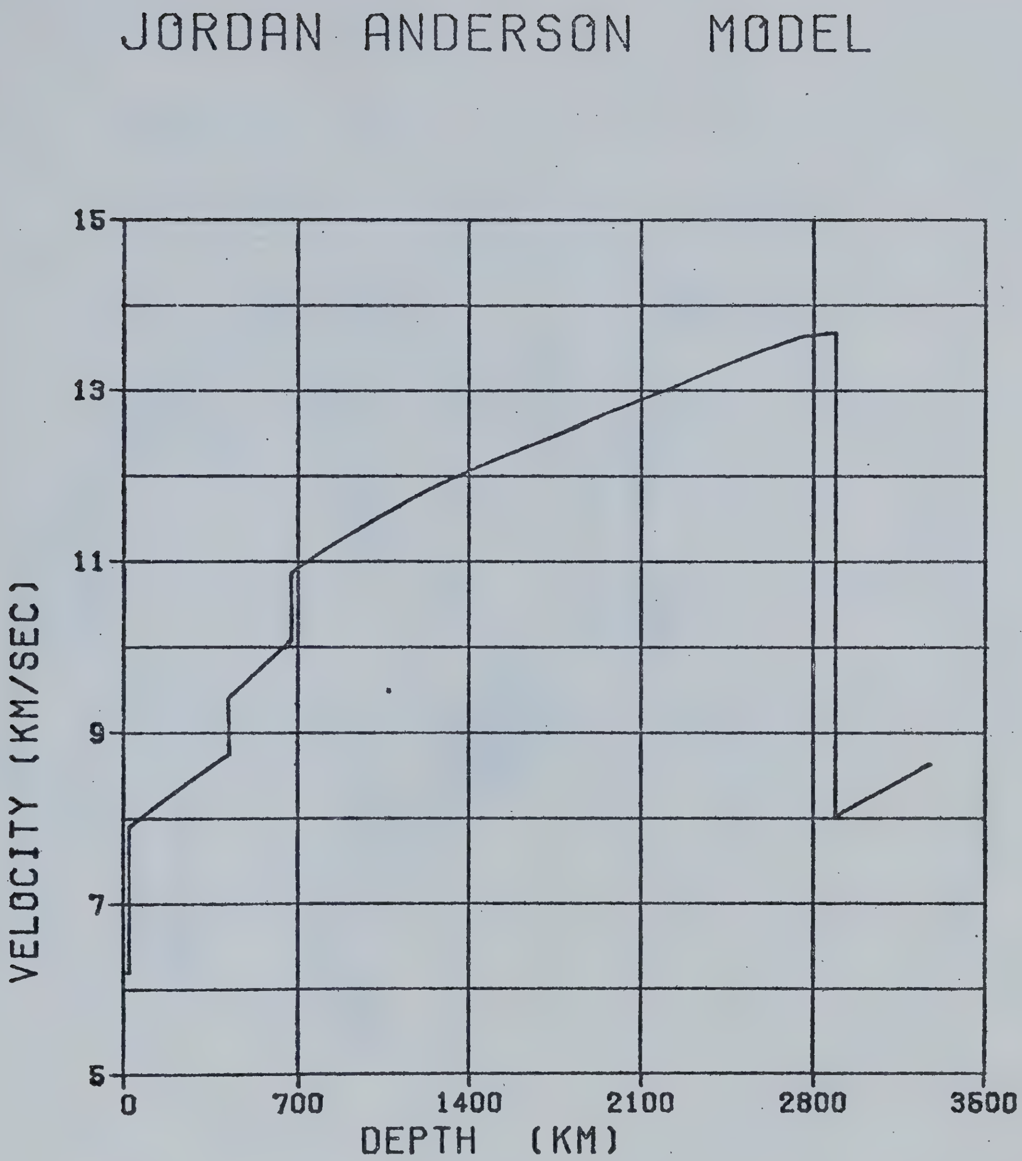

Figure 4-J : Model B1 (Jordan and Anderson, 1974) for P wave velocity of the mantle. This velocity model was inverted from free oscillation data and differential travel time data.

Figure 4-2 : This map shows the configuration of earthquakes (X) from the Philippine sources and receiving stations (⊡). The midpoints of the P rays (+) are located under the Kamchatka region.

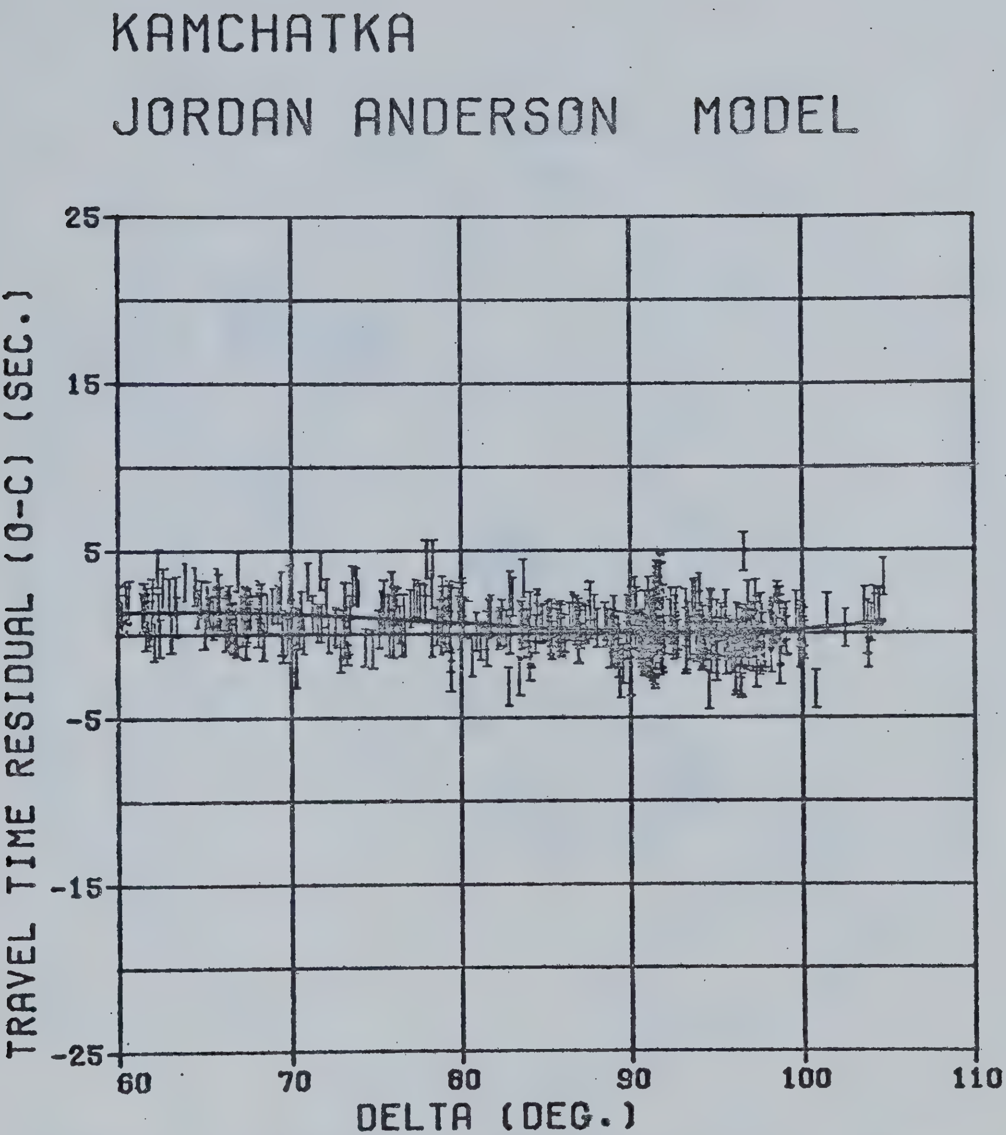

Figure 4-3 : Travel time residuals for the Philippine-Kamchatka ray path. The curve is a least-square fit cubic polynomial through the data points and the error bars are twice the standard deviation (1.1 sec.). There are 598 data points.

Figure 4-4 : Philippine- Mongolia ray path.

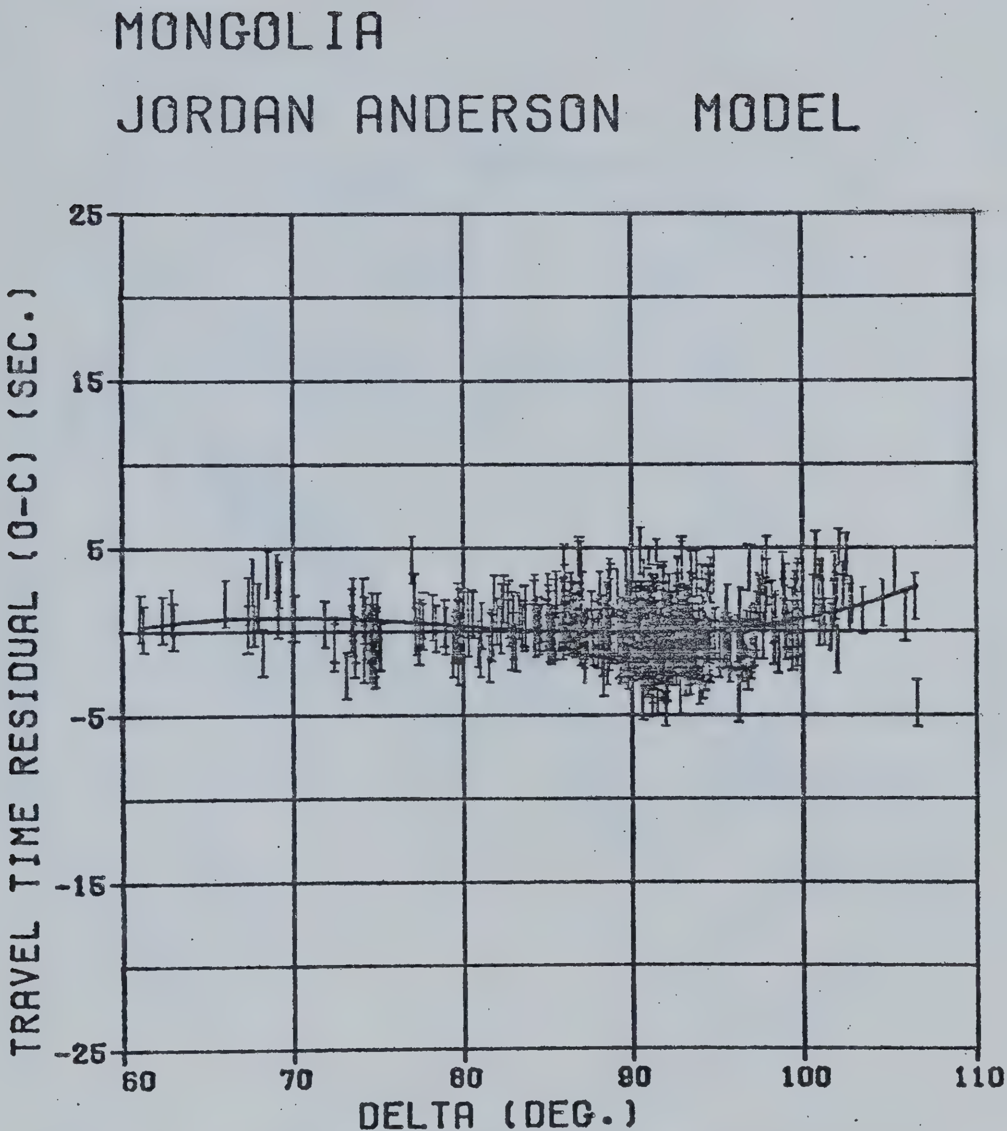

Figure 4-5 : Travel time residuals for the Philippine-Mongolia ray path. There are 769 points with standard deviation of 1.4 sec.

Figure 4-6 : Philippine-Himalaya ray path.

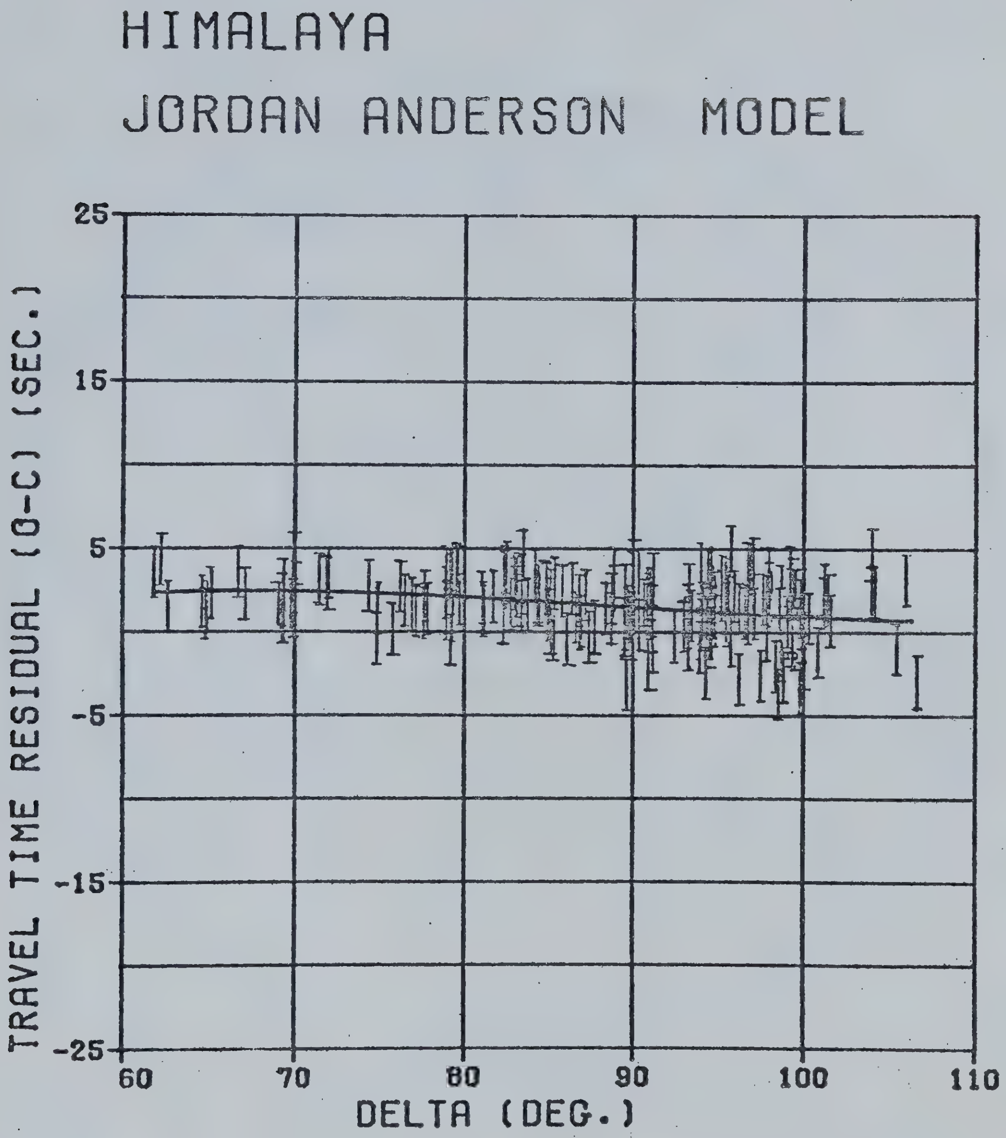

Figure 4-7 : Travel time residuals for the Philippine-Himalaya ray path. There are 212 points with standard deviation of 1.5 sec.

Figure 4-8 : Philippine-Indian Ocean ray path.

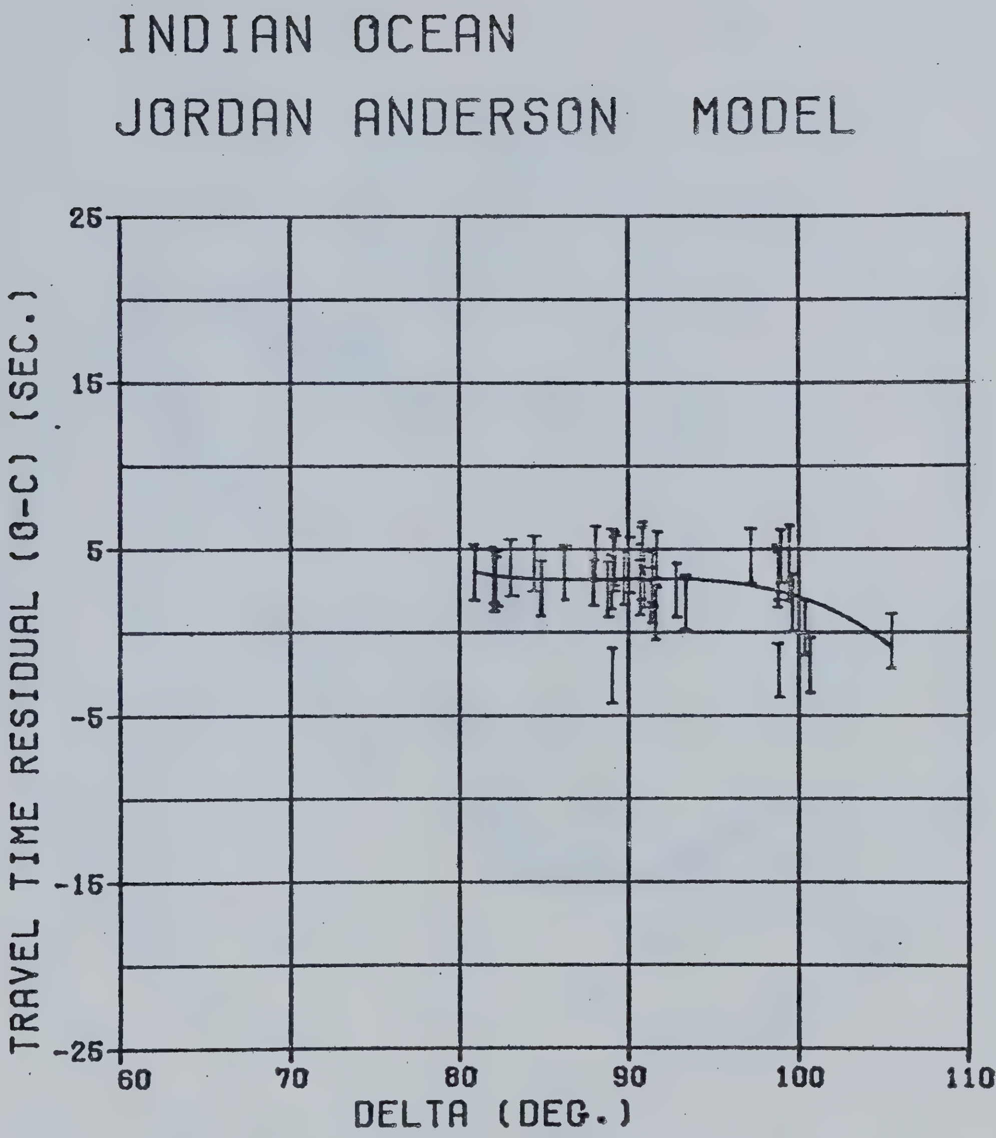

Figure 4-9 : Travel time residuals for the Philippine-Indian Ocean ray path. There are 46 points with standard deviation of 1.6 sec.

Figure 4-10 : Sunda-Mongolia ray path.

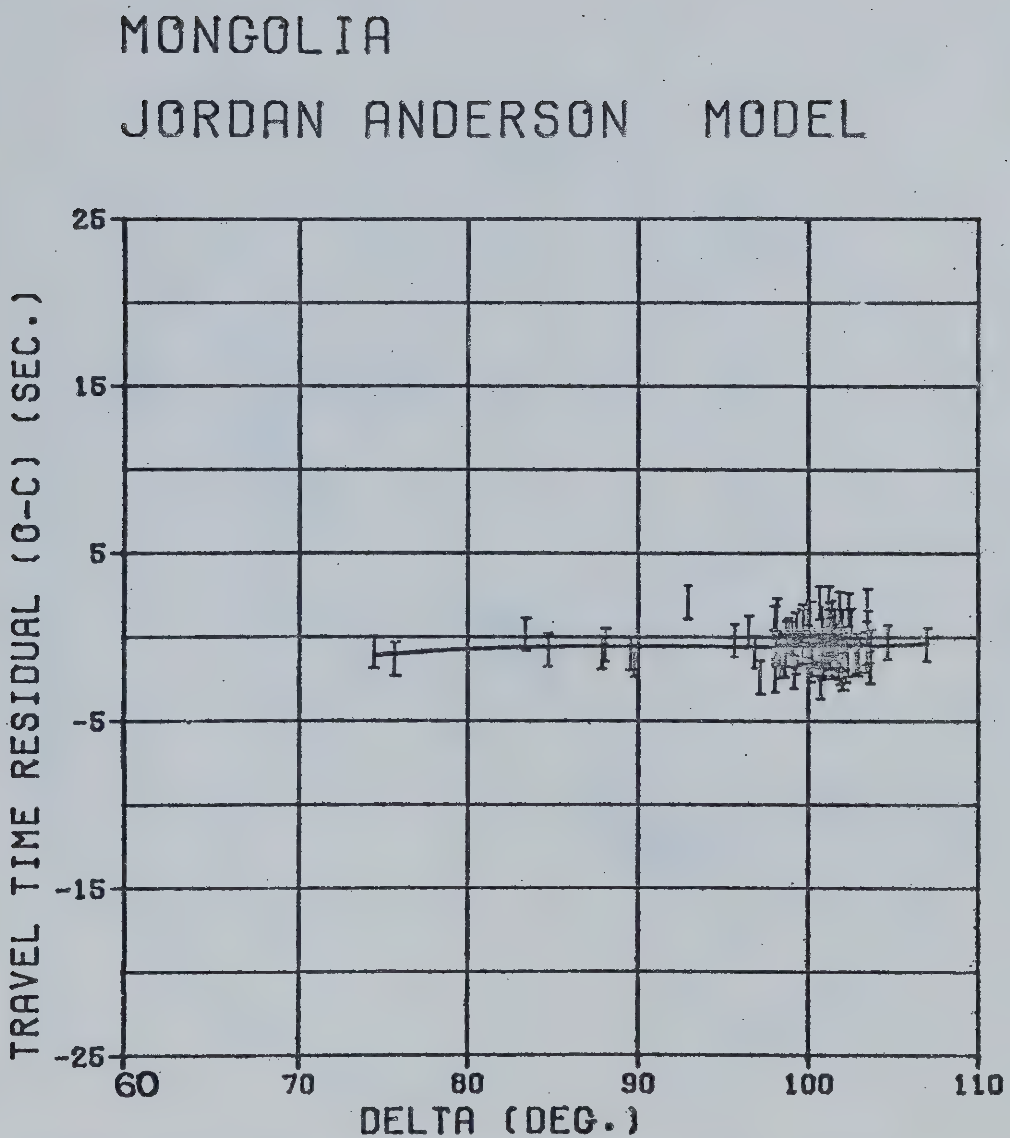

Figure 4-11 : Travel time residuals for the Sunda-Mongolia ray path. There are 119 points with standard deviation of 1.0 sec.

Figure 4-12 : Sunda-Himalaya ray path.

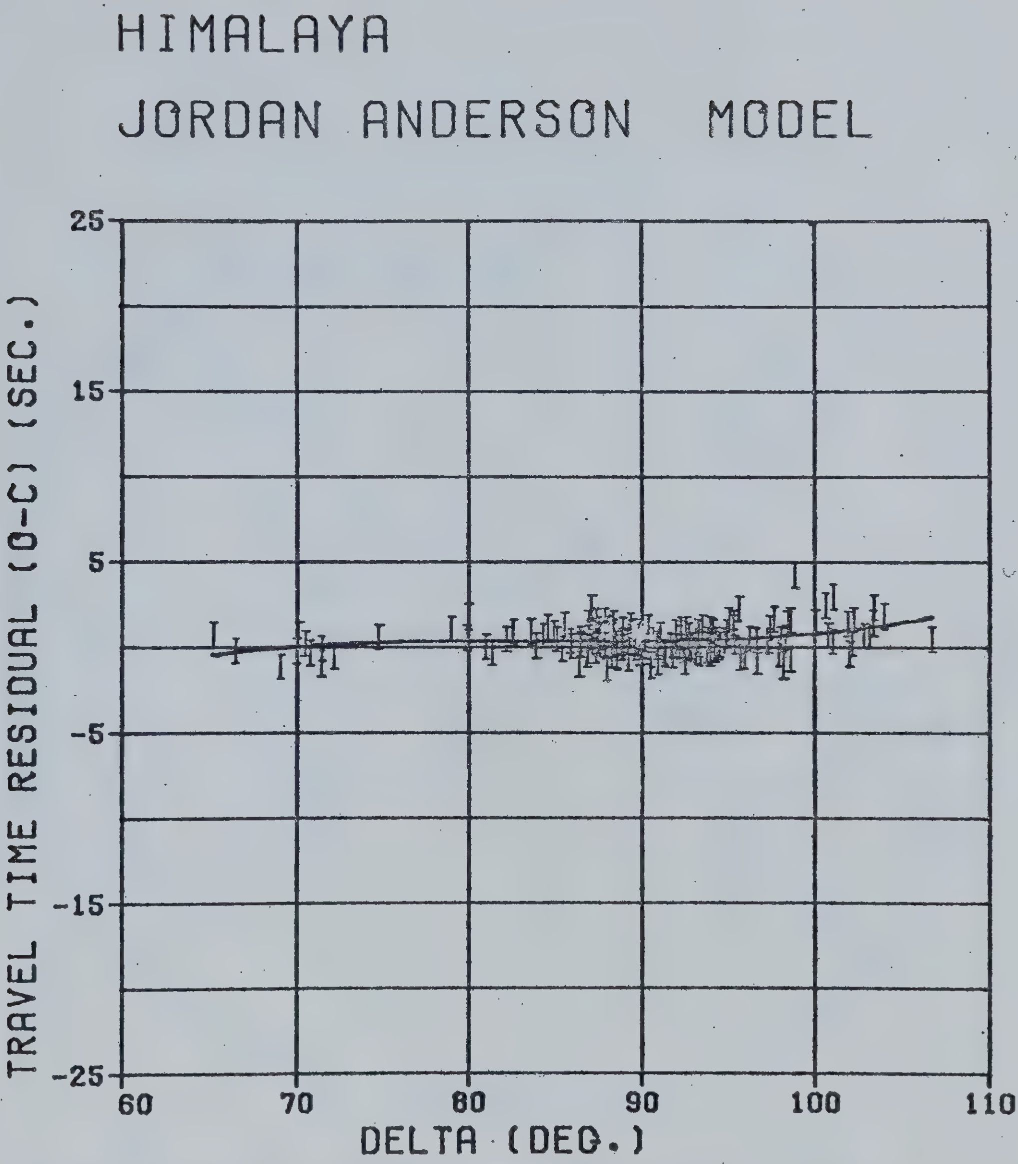

Figure 4-13 : Travel time residuals for the Sunda-Himalaya ray path. There are 182 points with standard deviation of 0.7 sec.

Figure 4-14 : Sunda-Indian Ocean ray path.

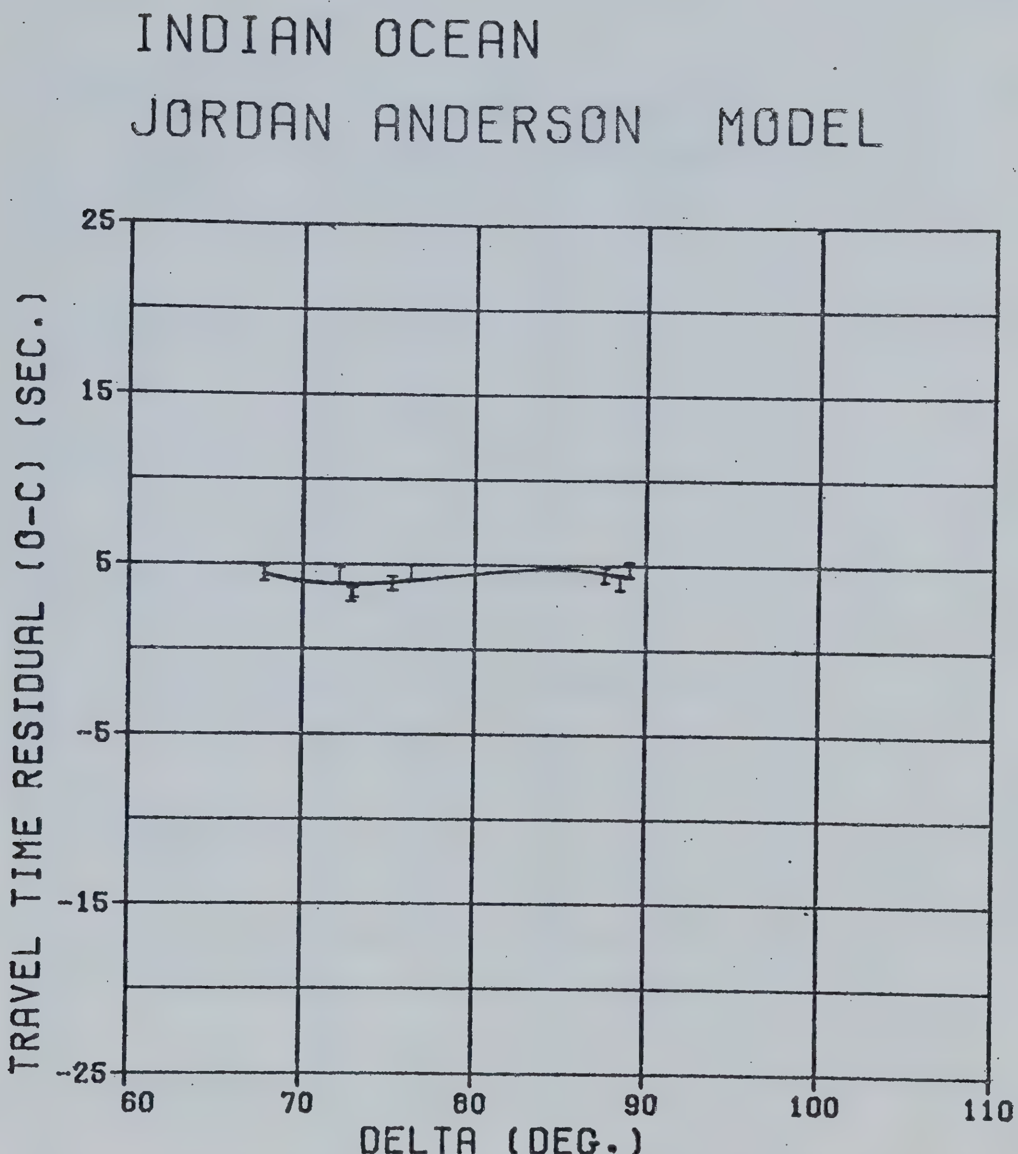

Figure 4-15 : Travel time residuals for the Sunda-Indian Ocean ray path. There are 9 points with standard deviation of 0.4 sec.

TABLE 4-1

REGION NAME	GEOGRAPHIC BOUNDARIES LATITUDE	LONGITUDE
KAMCHATKA	50 N	130 E
	70 N	155 E
	50 N	165 E
	43 N	148 E
MONGOLIA	33 N	102 E
	45 N	87 E
	52 N	100 E
	37 N	115 E
HIMALAYA	24 N	90 E
	30 N	75 E
	38 N	78 E
	32 N	95 E
INDIAN OCEAN	5 N	70 E
	10 N	70 E
	10 N	85 E
	5 N	85 E

TABLE 4-2

GRAPH OF	COEFFICIENT OF DETERMINATION			
	1ST	2ND	3RD	4TH
FIGURE 4-3	0.07	0.08	0.08	0.09
FIGURE 4-5	0.00	0.06	0.10	0.11
FIGURE 4-7	0.11	0.12	0.12	0.12
FIGURE 4-9	0.11	0.14	0.15	0.15
FIGURE 4-11	0.00	0.00	0.00	0.00
FIGURE 4-13	0.00	0.03	0.04	0.04
FIGURE 4-15	0.10	0.20	0.27	0.39

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